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## Exceptional atmospheric circulation during the “Dust Bowl”

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Fischer, Andreas M ; Schraner, Martin ; Peter, Thomas ; Rozanov, Eugene ; Ross, T

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# Exceptional atmospheric circulation during the “Dust Bowl”

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[1] The three-dimensional, regional and large-scale atmospheric circulation during the “Dust Bowl” is analyzed based on newly available historical upper-air data and reconstructed upper-level fields. The Great Plains Low Level Jet, transporting moisture into the region, was weakened on its eastern side, shallower, and penetrated less far north than during wet years. Nocturnal convection was likely suppressed by increased stability. Strong mid-tropospheric ridging was found over the Great Plains, and upper-tropospheric flow anomalies extended from the North Pacific across North America to the Atlantic. These findings provide a dynamical view of the “Dust Bowl” droughts, some aspects of which are distinct from other droughts. It is demonstrated that this is important for assessing predictive capabilities of current modeling systems. **Citation:** Brönnimann, S., A. Stickler, T. Griesser, T. Ewen, A. N. Grant, A. M. Fischer, M. Schraner, T. Peter, E. Rozanov, and T. Ross (2009), Exceptional atmospheric circulation during the “Dust Bowl,” *Geophys. Res. Lett.*, **36**, L08802, doi:10.1029/2009GL037612.

## 1. Introduction

[2] The midwestern United States is a region repeatedly affected by droughts, the most famous being the “Dust Bowl” droughts of the 1930s, with devastating social and economic consequences [Worster, 1979]. The mechanisms behind this event are still not fully understood. Based on model simulations, oceanic forcing has been suggested as a trigger [Seager *et al.*, 2005, 2008; Seager, 2007; Woodhouse and Overpeck, 1998; McCabe *et al.*, 2004; Schubert *et al.*, 2004a, 2004b; Cook *et al.*, 2008], amplified by land-atmosphere interactions [Seager *et al.*, 2005, 2008; Seager, 2007; Woodhouse and Overpeck, 1998; Schubert *et al.*, 2004a, 2004b] or atmospheric dust [Cook *et al.*, 2008]. However, due to lack of data, comparisons with observations were previously limited to the Earth’s surface (mostly precipitation), which is insufficient to conclusively address effects of remote forcings. Here we analyze the 3-dimensional atmospheric circulation during the “Dust Bowl” based on historical observations, reconstructions, and climate model simulations.

[3] The forcings responsible for the “Dust Bowl” droughts, as found in model simulations, are expected to proceed through changes in atmospheric circulation.

Therefore, causes are best identified by evaluating simulations with respect to the response of the atmospheric circulation to the imposed forcing. In the following we focus on three specific circulation features which based on the literature are expected to be critical to understanding the “Dust Bowl” droughts: the Great Plains Low Level Jet (GPLLJ), mid-tropospheric ridging, and the flow in the upper troposphere.

[4] The GPLLJ is responsible for most of the moisture advection from the Gulf of Mexico, the Caribbean Sea, and the North Atlantic into the Great Plains region [Rasmusson, 1967; Helfand and Schubert, 1995; Higgins *et al.*, 1997]. The GPLLJ also affects low-level convergence and nocturnal convective systems [Higgins *et al.*, 1997]. Interannual changes in the GPLLJ can be caused by a displacement or a change in strength of the climatological pressure centers, which in turn have been linked to sea-surface temperature (SST) anomalies in the tropical Pacific, to the North Atlantic Oscillation, and to the Asian Monsoon [Rodwell and Hoskins, 2001; Weaver and Nigam, 2008].

[5] Persistent mid-tropospheric ridging has been found to be a dominant meteorological feature during Great Plains droughts in the 1950s and later (when upper-air data are available) [Namias, 1982; Chang and Wallace, 1987]. Accompanying subsidence may suppress convection and thus reduce precipitation, which is predominantly convective in the Midwest during the warm season. Various factors such as large-scale oceanic forcing, regional land-atmosphere interaction, and atmospheric dust have been suggested to be involved in reinforcing the “Great Plains ridge”, as suggested by Namias [1982] in his visionary but observationally not well supported work.

[6] The flow in the upper troposphere also plays an important role. Tropical oceanic forcing likely results in changes in the Hadley circulation and upper tropospheric wave trains penetrating into the extratropics. The jet streams provide a waveguide for quasi-stationary waves and affect the propagation of rain-bearing disturbances from the Pacific into the continent in winter and spring. This might be important because soil moisture in spring has been suggested to affect summertime precipitation in the Great Plains [Oglesby, 1991; van der Schrier and Barkmeijer, 2007] and could also be of interest with respect to predictability.

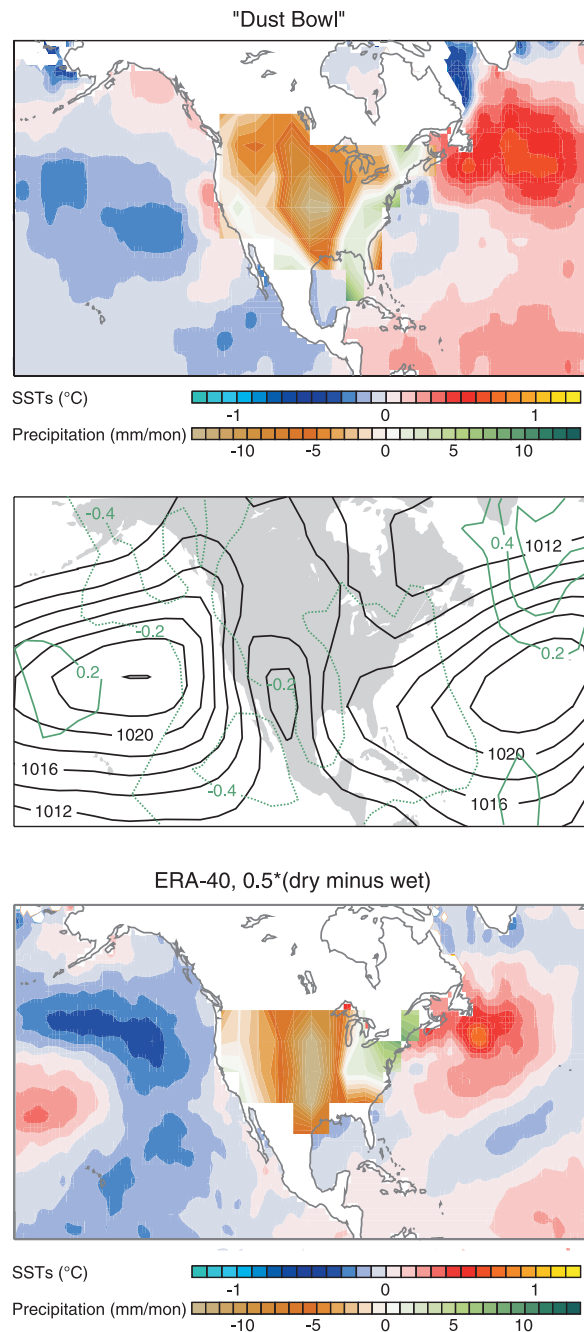
## 2. Data and Methods

[7] The historical upper-air data analyzed here include wind profiles obtained with pilot balloons (from the NOAA Climate Database Modernization Project, National Climatic Data Center) and temperature and pressure profiles from routine aircraft observations and (starting in 1938) radiosondes [Brönnimann, 2003; Ewen *et al.*, 2008]. Due to

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**Figure 1.** (top) AMJJA averages of precipitation and SST anomalies as well as (middle) SLP and SLP anomalies (hPa) for 1931–1939. Anomalies are with respect to 1921–1950. (bottom) Same as Figure 1 (top) but for dry minus wet years in ERA-40 data.

changes in the observing systems in the mid-1940s, we contrast data from the drought period (defined as 1931–1939) with data from adjacent years using the same observing system. The years 1941–1944 were wet in the Great Plains, with a precipitation anomaly almost exactly opposite to that in 1931–1939. The year 1940, which was relatively dry in the Great Plains, is considered a transitional year.

[8] In addition to observations we also present statistical reconstructions of upper-level geopotential height (GPH)

fields that are based on a large amount of global historical upper-air data, supplemented with land station temperature and sea-level pressure (SLP) data [Griesser *et al.*, 2008]. We express all fields as anomalies with respect to 1921–1950 to avoid effects of long-term trends. The ERA-40 reanalysis [Uppala *et al.*, 2005] is used as a counterpart in the more recent past.

### 3. Results and Discussion

#### 3.1. Surface Fields

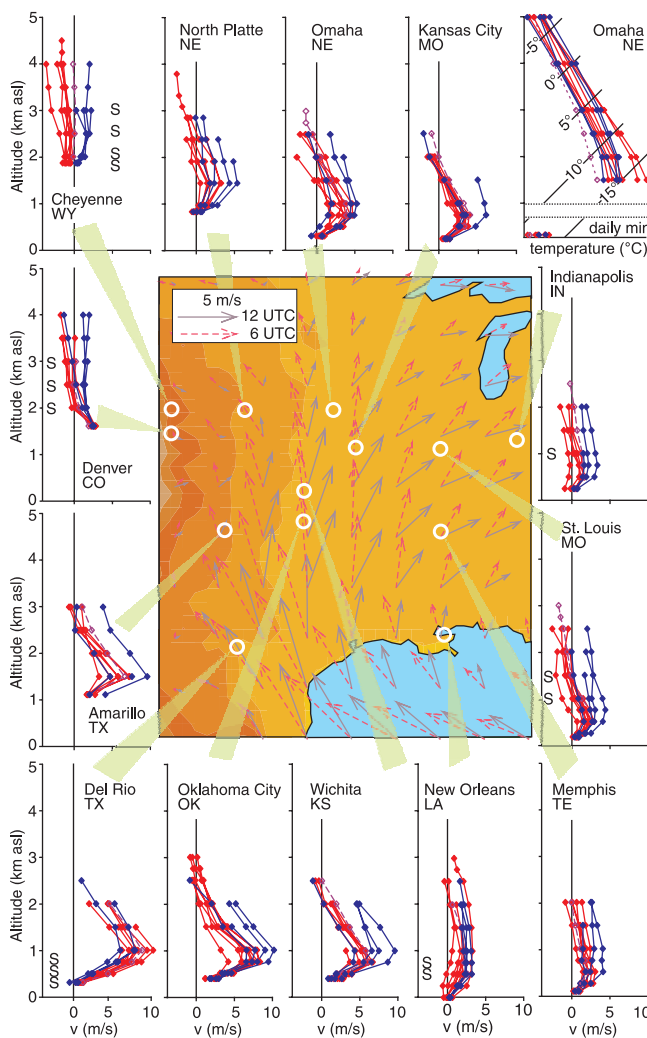
[9] Observed precipitation (GHCnv2) anomalies [Vose *et al.*, 1992] from April to August (AMJJA) for 1931–1939 are shown in Figure 1 together with corresponding SST anomalies (HadISST2) [Rayner *et al.*, 2006], and SLP (HadSLP2) [Allan and Ansell, 2006]. Precipitation was strongly reduced in the Great Plains region (but arguably increased over the Gulf of Mexico). As is well known from other studies, the North Pacific was anomalously cold while the North Atlantic was warm. The two dominating high pressure systems, the North Pacific and Azores-Bermuda highs, show slight westward and northeastward shifts, respectively. Without wind data, however, no conclusions can be drawn about atmospheric circulation changes.

#### 3.2. Great Plains Low-Level Jet

[10] Because the GPLLJ changes during the course of a night (Figure 2, middle), we show only late night (3:00–6:00 local time) measurements ( $\geq 18$  profiles must be available per month). Figure 2 shows the averaged AMJJA meridional wind profiles for each year for selected sites. Despite the limited vertical resolution, the low-level jet structure is clearly visible. During the drought period (red), the meridional component of the GPLLJ was generally weaker than during the early 1940s (blue). However, this does not hold for the core of the jet (e.g., Del Rio, TX). The difference was largest above the altitude of the jet maximum (e.g., Oklahoma City, OK; Wichita, KS) and east of the jet core (e.g., St. Louis, MO). Qualitatively similar results were also found for the early night (21:00–24:00) and for the daytime (but with a weaker jet; results not shown) and hence can be considered robust.

[11] Wind fields at 1000 and 2000 m asl (Figure 3) further confirm that the main signal is not the weakening of the average jet structure but a change in orientation (stronger westerly component) and northward extent. In this respect the “Dust Bowl” differs from other droughts [e.g., Lyon and Dole, 1995]. In fact, an analysis of nocturnal (12 UTC) wind fields for dry minus wet summers in ERA-40 (defined as AMJJA precipitation anomalies in the Great Plains, as those of Schubert *et al.* [2004a], outside  $\pm 8$  mm/month) clearly shows a weakening of the GPLLJ in its core region (Figure 3, top right). Other aspects of the “Dust Bowl” are similar to recent droughts, e.g., that the signal is largest above the jet maximum.

[12] The GPLLJ change between the “Dust Bowl” and the early 1940s shows weak intra-seasonal dependence, with a slightly stronger signal in summer than in spring. In contrast, in the ERA-40 analysis the largest change is found in spring. Furthermore, almost only westerly wind anomalies (compared to 1941–1944) were found during the “Dust Bowl” years, but easterly anomalies in the ERA-40 analysis in the northern regions in spring and in the southern



**Figure 2.** Nocturnal meridional wind profiles (AMJJA averages) over the Great Plains for 1931–1939 (red) and 1941–1944 (blue, significant differences between the two (heteroskedastic t-test,  $\alpha = 0.05$ ) are marked with “S”). Profiles for 1940 are shown in pink. (middle) AMJJA winds at 925 hPa from ERA-40 (1961–1990) at 6 UTC (dashed) and 12 UTC (solid) for orientation. (top right) AMJJA averaged nocturnal temperature profiles from Omaha, NE (1932–1944) and daily minimum temperatures at Tekamah, NE.

regions in summer. The differences between the two analyses are largest in summer.

[13] In summary, wind observations show that during the “Dust Bowl” years, the GPLLJ was weaker on its eastward side, shallower, and penetrated less far north in comparison to the early 1940s. Humidity data are not available for the 1930s. However, other studies show that specific humidity, on an interannual scale is less important than wind speed for moisture transport through the GPLLJ [Wang *et al.*, 2008]. This suggests weakened moisture transport induced by the weakened GPLLJ during the “Dust Bowl” droughts.

### 3.3. Thermal Structure and Mid-tropospheric Ridging

[14] Since warm season precipitation is primarily convective, we analyzed temperature profiles from nocturnal

aircraft and radiosonde ascents from Omaha, NE, the only continuous record in the region, in order to assess the thermal structure of the atmosphere. Observation times changed in several steps from 10 to 4 UTC. We adjusted the data to 6 UTC based on interpolated 6-hourly climatologies from ERA-interim (1989–2005), however, results should be interpreted with care. The two hot and dry summers 1934 and 1936 clearly stand out (Figure 2, top right) with a maximum warming at 1.5 km asl.

[15] Because surface data could not be adjusted, we analyzed daily minimum temperatures at nearby Tekamah. In contrast to the daily maxima, the minima were only slightly higher in 1934 and 1936 (and even lower in the 1931–1939 average) than in 1941–1944. Even though daily minima cannot substitute nocturnal temperature, the contrast to the situation at 1.5 km asl is very large and implies stronger stability in the lowest kilometer and possibly suppressed convection during the night, likely accentuated by decreased low-level humidity (note that stability was decreased above 1.5 km asl).

[16] Large-scale subsidence might have contributed to the thermal structure. The reconstructed 500 hPa GPH fields (Figure 4) show ridging in 1931–1939, with the largest positive anomalies northeast of the Great Plains and additional centers in the Gulf of Alaska and over the Atlantic. Radiative effects of atmospheric dust or feedbacks involving the land surface could have contributed to the ridging and the vertical thermal structure over the Great Plains, but need to be assessed by detailed comparisons with targeted model simulations.

### 3.4. Large-Scale Upper-tropospheric Flow

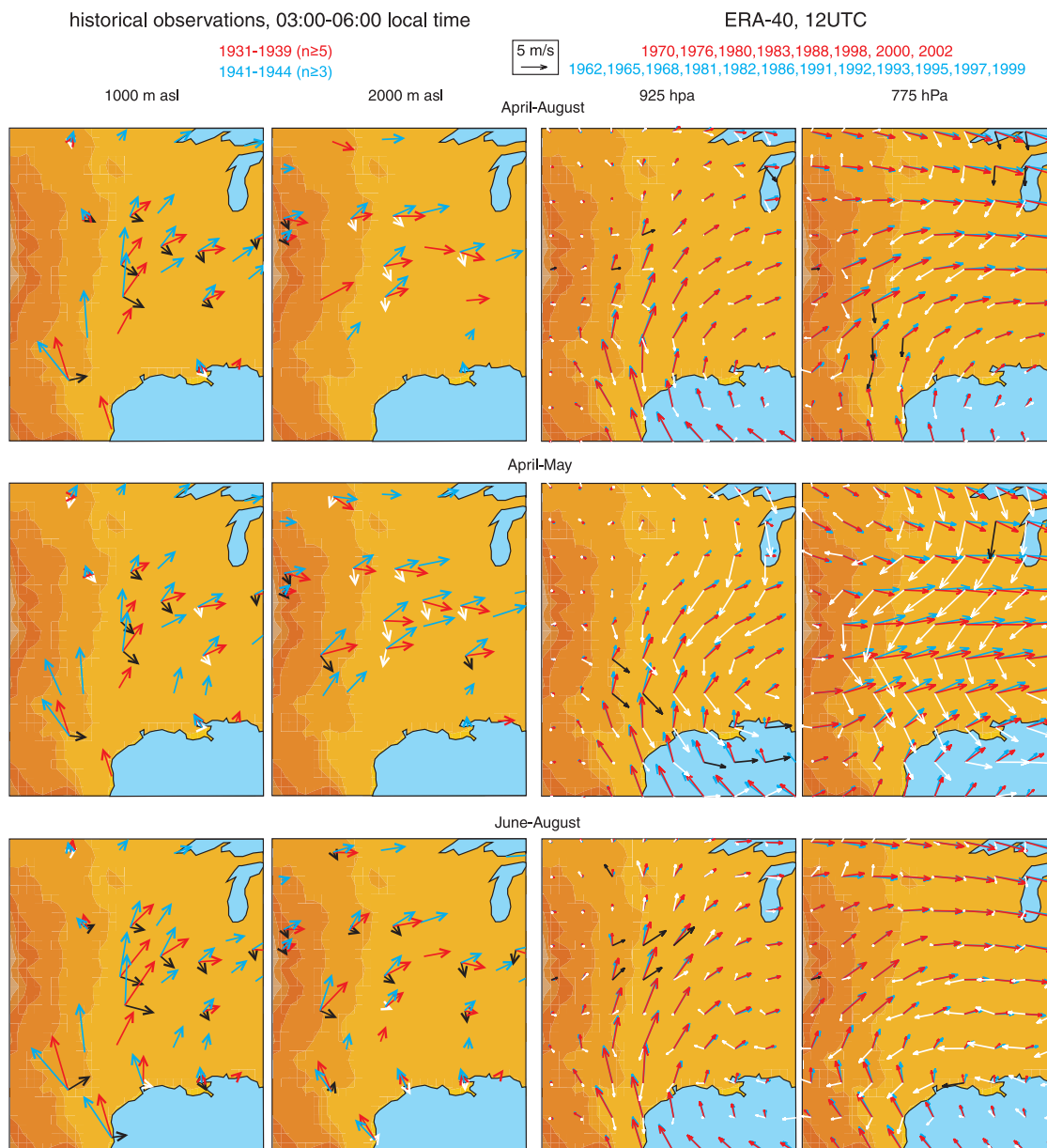
[17] Our reconstructed 200 hPa GPH fields indicate persistent changes in the upper tropospheric mean flow (Figure 4). The averaged winter field (Nov–Mar preceding the analyzed summer seasons) shows a zonally structured, positive GPH anomaly stretching from the North Pacific into the North Atlantic (with three centers) and a strong negative anomaly over northwestern Canada. The effect of tropical Pacific forcing on the extratropics is strongest in winter and is likely reflected in this pattern. In spring, positive anomalies off the coast of California might indicate weaker Pacific jets and possibly fewer disturbances that would otherwise bring in moisture from the Pacific (note that neither reconstructions nor observations allow addressing jet streams and disturbances directly). In summer, a positive GPH anomaly centre is located over the continent. GPH gradients suggest a poleward migration of the polar front jet over the Great Plains, which is consistent with (albeit sparse) upper tropospheric wind observations from the northern plains.

[18] Note that in all seasons, the largest GPH anomaly is found over northeastern Europe. Though an influence on the Great Plains is unlikely, this anomaly might have played a role for the concurrent extremely warm years in the European Arctic.

### 3.5. Comparison With Model Simulations

[19] The identified characteristic flow features can now be used to assess SST-forced climate model simulations. In addition to published studies, we also refer to our own simulations with the SOCOL model [Schraner *et al.*, 2008;





**Figure 3.** Nocturnal winds at two levels averaged for different seasons for drought years (red arrows) and wet years (blue arrows; white arrows indicate the difference). (left) Pilot balloon data from 1931–1939 and 1941–1944. (right) A corresponding analysis using ERA-40 data, with difference vectors multiplied by 5. Difference vectors whose  $u$  or  $v$  component was statistically significant (heteroskedastic  $t$ -test,  $\alpha = 0.05$ ) are shown in black.

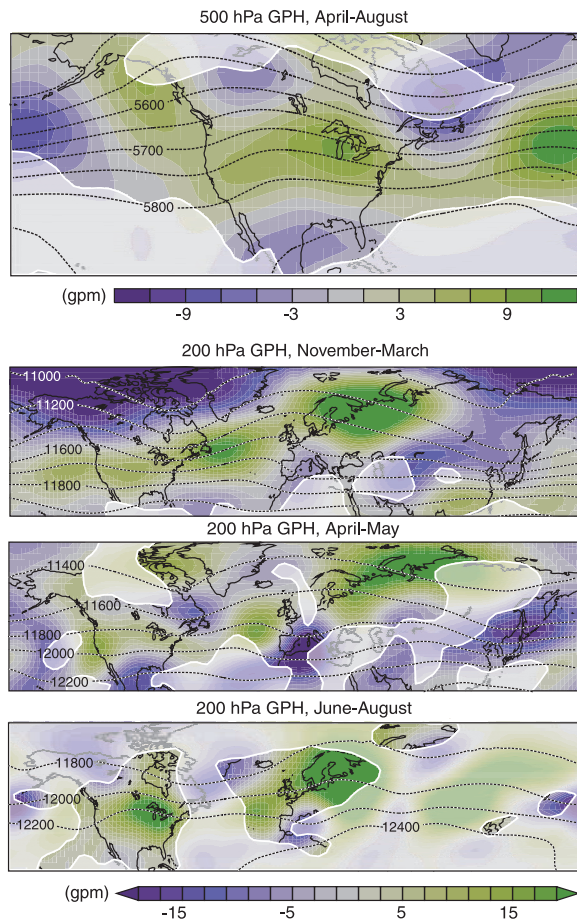
*Fischer et al.*, 2008] (see the auxiliary material) for illustrative purposes.<sup>1</sup>

[20] Many models, including Socol (Figure S1), reproduce the drought in the Midwest in the 1930s in the ensemble mean when forced with observed SSTs and also show increased precipitation over the Caribbean. However, many also produce strong precipitation deficits in Northern Mexico, which was not observed [*Seager et al.*, 2005, 2008; *Schubert et al.*, 2004a; *Cook et al.*, 2008]. Socol does this in each ensemble member (Figure S1), which could point to a misrepresentation of the regional circulation.

<sup>1</sup>Auxiliary materials are available in the HTML. doi:10.1029/2009GL037612.

[21] The large-scale circulation response is reproduced reasonably well in the models. 200 hPa GPH patterns in published studies are in good agreement with our reconstructions [*Seager et al.*, 2005; see also *Schubert et al.*, 2004b], thus lending credibility to the models. Socol has the band of positive GPH anomalies across the Pacific North American sector too far south in the ensemble mean (Figure S3, left), but the member with the strongest response in Great Plains precipitation also shows the best agreement in 200 hPa GPH (Figure S3, right).

[22] Most studies do not show modeled wind fields, but results from *Seager et al.* [2005] and Socol (Figure S2) agree better with the ERA-40 analysis than with the “Dust Bowl” as they show easterly wind anomalies and suggest a



**Figure 4.** Reconstructed anomalies (with respect to 1921–1950) of 200 and 500 hPa GPH for different seasons from 1931–1939. Dashed lines give absolute values (gpm). Whitened areas denote low reconstruction skill (reduction of error  $<0.3$  [see Griesser et al., 2008]).

weakening of the jet core. A limitation in reproducing regional circulation changes evidently might affect the simulated spatial pattern of precipitation.

#### 4. Discussion and Conclusions

[23] The historical upper-air data and reconstructions provide a dynamical view of the “Dust Bowl” droughts (a conceptual depiction is shown in Figure S4), which is necessary to understand its causes. The trigger undoubtedly was oceanic forcing. Only SST-forced model simulations reproduce the drought tendency in the Great Plains as well as the large-scale flow in the upper-troposphere. This suggests that there is predictability (to the extent to which SSTs are predictable) in the likelihood of “Dust Bowl”-like droughts occurring in North America.

[24] What is the relative role of Pacific and Atlantic SSTs? Pacific SSTs affect the upper-level circulation year-round and the GPLLJ in spring and summer while Atlantic SSTs have their largest influence in summer and fall and also affect the GPLLJ [e.g., Schubert et al., 2004a; Wang et al., 2008; Weaver and Nigam, 2008]. The upper-level anomalies and the unanimous weakening in the GPLLJ

core in spring could therefore reflect Pacific influences, while the summer signal during the “Dust Bowl” (which is distinct from more recent droughts) could point to an additional Atlantic influence. In fact, Caribbean SST anomalies were warmer during the “Dust Bowl” compared to other droughts (Figure 1). The causes for the mismatch in the spatial pattern of precipitation and wind anomalies between the “Dust Bowl”, recent droughts, and model simulations need to be further studied.

[25] The results presented in this paper show that the historical data and reconstructions can serve as a benchmark for climate model evaluation. They allow a refined, process-based assessment of climate models ranging from regional to large-scale circulation responses.

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